

**RESEARCH LETTER**

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**Key Points:**

- Proxy-based reconstructions and model-based simulations of global mean surface temperature over the last 800,000 years differ in detail
- During periods of decreasing obliquity and sea level the proxy reconstructions show a temperature- $\text{CO}_2$  divergence missing in simulations
- Elimination of these periods leads to a more linear paleoclimate sensitivity and to equilibrium warming for  $\text{CO}_2$  doubling of 2–4 K

**Supporting Information:**

- Supporting Information S1

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# The Effect of Obliquity-Driven Changes on Paleoclimate Sensitivity During the Late Pleistocene

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**Abstract** We reanalyze existing paleodata of global mean surface temperature  $\Delta T_g$  and radiative forcing  $\Delta R$  of  $\text{CO}_2$  and land ice albedo for the last 800,000 years to show that a state-dependency in paleoclimate sensitivity  $S$ , as previously suggested, is only found if  $\Delta T_g$  is based on reconstructions, and not when  $\Delta T_g$  is based on model simulations. Furthermore, during times of decreasing obliquity (periods of land ice sheet growth and sea level fall) the multimillennial component of reconstructed  $\Delta T_g$  diverges from  $\text{CO}_2$ , while in simulations both variables vary more synchronously, suggesting that the differences during these times are due to relatively low rates of simulated land ice growth and associated cooling. To produce a reconstruction-based extrapolation of  $S$  for the future, we exclude intervals with strong  $\Delta T_g$ - $\text{CO}_2$  divergence and find that  $S$  is less state-dependent, or even constant state-independent), yielding a mean equilibrium warming of 2–4 K for a doubling of  $\text{CO}_2$ .

**Plain Language Summary** Anthropogenic carbon dioxide ( $\text{CO}_2$ ) emissions will lead to rising global mean temperature through the greenhouse effect. The amplitude of this warming, as estimated with computer simulations for the equilibrium climate response to a doubling of atmospheric  $\text{CO}_2$  concentration, is called climate sensitivity. It is necessary to verify these simulation-based quantifications of climate sensitivity with independent alternative approaches. One such approach is the analysis of past (paleo) climates, which has indicated a state-dependent paleoclimate sensitivity. Here we compare different data-based reconstructions and computer-based simulations of paleoclimate sensitivity of the last 800,000 years and find that they disagree. In data-based reconstructions global mean temperature and  $\text{CO}_2$  diverge during intervals when land ice growth is particularly pronounced. This temperature- $\text{CO}_2$  divergence is not observed in simulations, probably due to an underestimation of the rate of land ice growth and the associated cooling. However, these periods of pronounced land ice growth are not of relevance for a warming future and can therefore be neglected when estimating climate sensitivity from reconstructions of the past. Consequently, we find that paleoclimate sensitivity derived from reconstructions is less state-dependent than previously thought and agrees with warming estimates of 2–4 °C as derived from simulated equilibrium climate response for  $\text{CO}_2$  doubling.

## 1. Introduction

Analyses of paleoreconstructions (Köhler et al., 2015; K2015 in the following) and paleoclimate simulations (Friedrich et al., 2016; F2016 in the following), covering the late Pleistocene, have suggested that climate sensitivity might not be a constant parameter of the climate system but a state-dependent variable that increases toward warmer climates. Most other studies on this topic indicate a similar behavior, including a review that covers a wide range of colder and warmer climate states (von der Heydt et al., 2016). However, there have also been studies using general circulation models (GCMs) or Earth system models of intermediate complexity (EMICs) which simulate an increase in climate sensitivity for colder than present-day climate (e.g., Colman & McAvaney, 2009; Kutzbach et al., 2013; Pfister & Stocker, 2017).

Fueled by this ambiguity we wanted to test the robustness of the conclusions in earlier studies (K2015 and F2016). Here we investigate whether this, previously found, state-dependency of climate sensitivity can be

reproduced in other setups; we reanalyze the proxy-based reconstructions of global temperature change ( $\Delta T_g$ ) published in the last few years (Snyder, 2016, in addition to K2015 and F2016), investigate transient 800-kyr simulation results obtained with the EMICs, CLIMBER (Ganopolski & Calov, 2011), and LOVECLIM (F2016), and analyze the only available transient GCM simulation across the last glacial/interglacial transition provided by the CCSM3 model (He, 2011; Liu et al., 2009; Figure 1).

A direct comparison of today's anthropogenic warming with paleodata-based reconstructions is not possible, due to the lack of a direct analog in the magnitude of the rate of changes. However, we can evaluate the general climate system response to radiative forcing anomalies. For such efforts, the specific equilibrium climate sensitivity (ECS)  $S_{[X]}$  (or paleoclimate sensitivity) has been defined as the ratio of the global and annual mean surface temperature change ( $\Delta T_g$ ) over the change in radiative forcing ( $\Delta R_{[X]}$ ) caused by the process(es) X (PALAEOSSENS-Project Members, 2012)

$$S_{[X]} = \frac{\Delta T_g}{\Delta R_{[X]}} \quad (1)$$

Here we calculate radiative forcing for processes including the greenhouse gas (GHG) effect ( $\text{CO}_2$ ,  $\text{CH}_4$ , and  $\text{N}_2\text{O}$ ) but also other processes, such as the (planetary) albedo effects from land ice (LI), vegetation (VG), and aerosols (AE). The time dependency of the climate to those forcing or feedback processes is not of particular interest in the following but has been addressed elsewhere (e.g., Rohling et al., 2018; Zeebe, 2013). This concept of calculating  $S_{[X]}$  was introduced in PALAEOSSENS-Project Members (2012) to clarify which forcing is explicitly included when estimating climate sensitivity from paleodata, not to test causation. Furthermore, this approach assumes that different forcing processes have a similar impact on  $\Delta T_g$ , which is a simplification (e.g., Stap et al., 2018; Yoshimori et al., 2011), that is difficult to overcome in analyses of mainly proxy-based reconstructions. Within the context of Earth system model analysis this ratio  $\Delta T_g/\Delta R_{[X]}$  is also called the *climate sensitivity parameter* (e.g., Yoshimori et al., 2011).

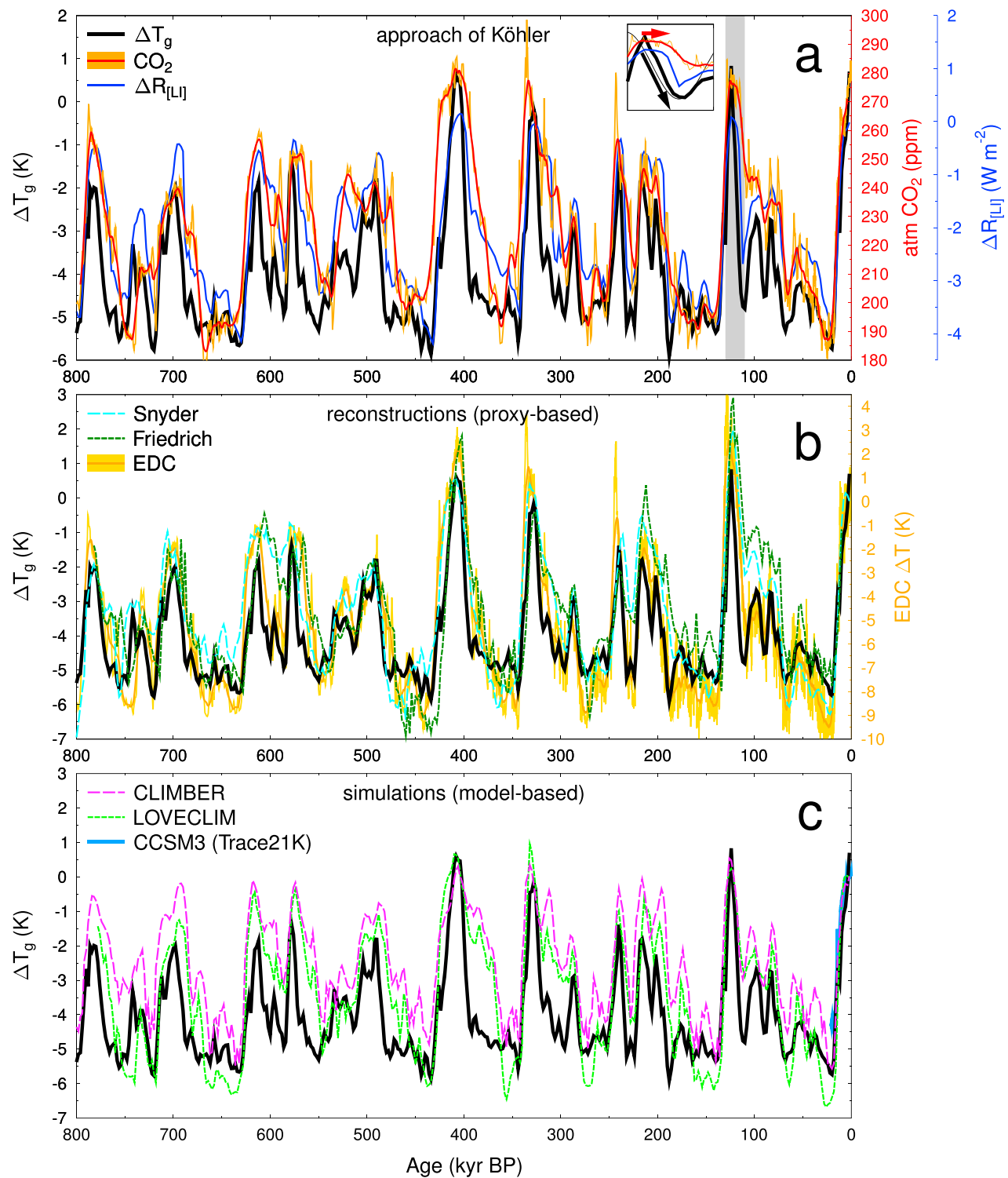
The emergence of state-dependency in  $S_{[X]}$  implies that the best fit to a scatter plot of  $\Delta T_g$  versus  $\Delta R_{[X]}$  is not linear, but some nonlinear function, for example, a higher-order polynomial (Figure 2a). While the detection of such a nonlinearity is rather straightforward, the quantification of  $S_{[X]}$  is more complicated, as described in detail by Köhler, Stap, et al. (2017).

In F2016 two independent estimates of  $\Delta T_g$  were generated: a purely proxy-based reconstruction based on SST data from 63 records and a simulation with the LOVECLIM model. The estimates of  $\Delta T_g$  were then averaged and confirmed the state-dependency in  $S_{[X]}$  for the last ~800 kyr as deduced by K2015. Since this state-dependency in  $S_{[X]}$  suggests that during warm interglacials a relatively small change in  $\Delta R$  leads to a relatively large change in  $\Delta T_g$  (Figure 2a), it is crucial to know how robust this conclusion is. Recently, a new proxy-based reconstruction of global mean temperature changes constructed from 61 records of SST anomalies has been published (Snyder, 2016). These two proxy-based reconstructions of  $\Delta T_g$  (F2016; Snyder, 2016) are not fully independent with respect to the underlying data but differ in details and in the upscaling methodologies.

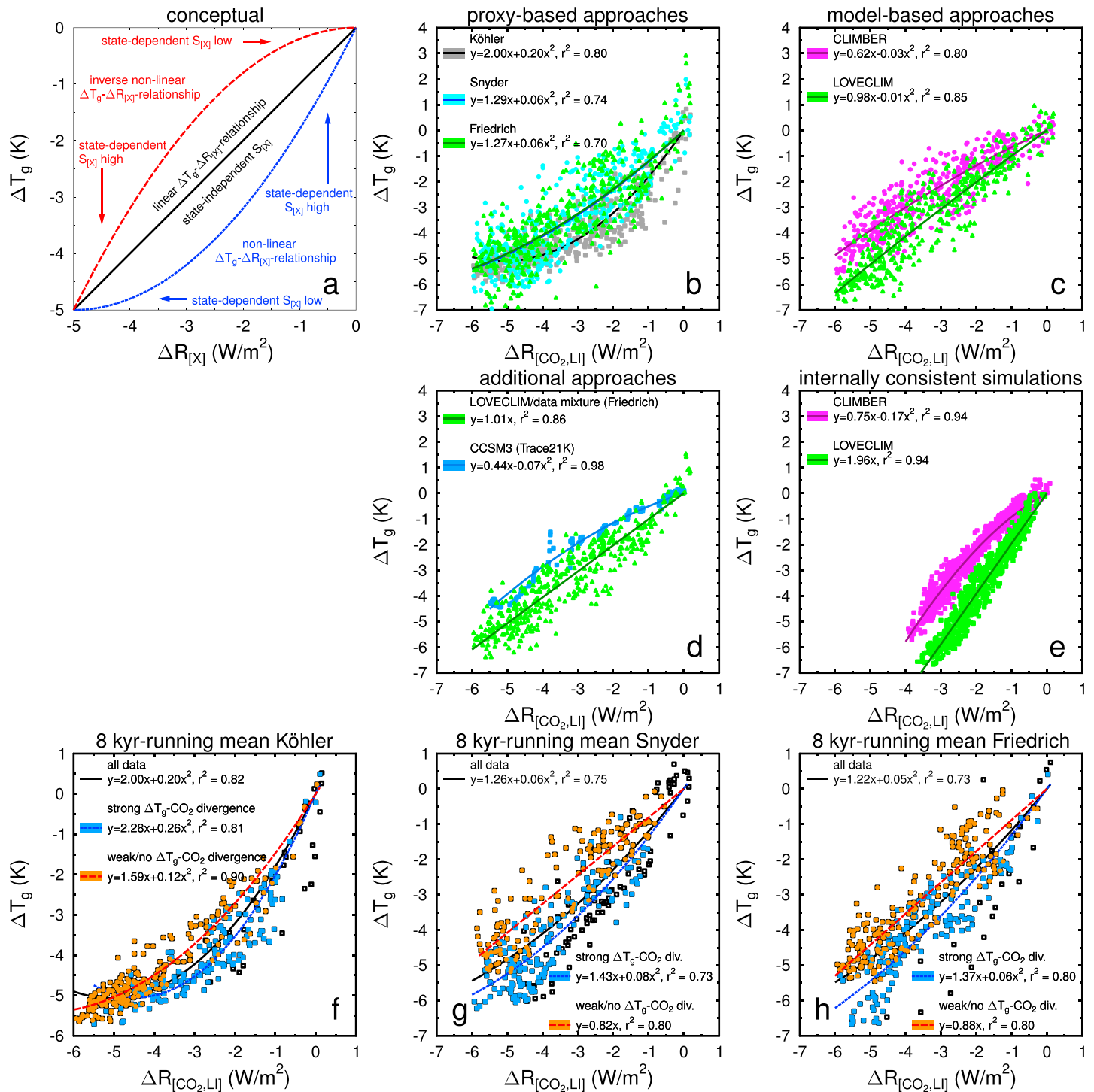
Finally, we discuss how our findings for paleoclimate sensitivity can be extrapolated to the future and compare a rough approximation of equilibrium global warming caused by  $2\times\text{CO}_2$  with other approaches.

## 2. Methods

In K2015 deconvolution of the LR04 benthic  $\delta^{18}\text{O}$  stack (Lisiecki & Raymo, 2005) was used to provide mutually consistent contributions from sea level (or land ice volume) and deep ocean temperature ( $\Delta T_o$ ) using 3-D ice sheet models of de Boer et al. (2014). Temperature change over land in the high-latitude Northern Hemisphere (about  $40\text{--}85^\circ\text{N}$ ,  $\Delta T_{\text{NH}}$ ) where most glacial/interglacial changes in land ice occurred during the late Pleistocene is linearly related to  $\Delta T_o$  on a multimillennial time scale. However,  $\Delta T_{\text{NH}}$  also contains changes due to elevation changes (lapse rate) and considers seasonality.  $\Delta T_g$  and  $\Delta T_{\text{NH}}$  are then related to each other via a nonconstant polar amplification factor ( $f_{\text{pa}}$ ) that has been determined from PMIP3 output. Sensitivity analyses (de Boer et al., 2014; K2015) have shown that  $\Delta T_g$  has a relative uncertainty of ~10% over the last 800 kyr. This setup is a model-based interpretation of proxy data. It is a mixture between a purely proxy-based reconstruction and model-based simulations. However, while full climate models are driven by temporal changes in various boundary conditions (e.g., insolation and GHG) and then calculate all other variables internally, here only the ice sheet dynamics are simulated. Therefore, we consider our approach to be more similar to those of



**Figure 1.** Paleodata of the last 800 kyr. (a) Data used in the approach of Köhler (K2015) with global mean temperature change  $\Delta T_g$ , land ice-based radiative forcing change  $\Delta R_{[LI]}$  and atmospheric  $\text{CO}_2$  (Bereiter et al., 2015). Inset shows an enlarged view on the divergence of  $\Delta T_g$  and  $\text{CO}_2$  at the end of the Eemian (130–110 kyr BP, gray band), including as thin black line changes in obliquity (Laskar et al., 2004). Comparing different temperature time series with K2015- $\Delta T_g$  (black bold line); (b) proxy-based reconstructions of  $\Delta T_g$  (Synder, Friedrich (F2016)) and EPICA Dome C (EDC)  $\Delta T$ ; (c) model-based simulations (CLIMBER and LOVECLIM) including CCSM3 for the last 21 kyr. Ice core data (EDC  $\Delta T$ ,  $\text{CO}_2$ ) are plotted on the most recent age model AICC2012 (Bazin et al., 2013; Veres et al., 2013) and shown as original high-resolution data (thin) and 8-kyr running means (bold).



**Figure 2.** Scatter plots of temperature change  $\Delta T_g$  over radiative forcing change  $\Delta R_{[X]}$ . (a) Conceptual understanding of different relationships between  $\Delta T_g$  and  $\Delta R_{[X]}$  and the resulting state-(in)dependency of  $S_{px}$ . (b) Data-based reconstructions of  $\Delta T_g$  (Köhler, Snyder, and Friedrich); (c) model simulation results of  $\Delta T_g$  (CLIMBER and LOVECLIM); (d) alternative approaches (Friedrich's model/data mixture for  $\Delta T_g$ , 21 kyr transient simulations with CCSM); (e) internally consistent model setups of CLIMBER and LOVECLIM; (f–h) multimillennial component (8-kyr running mean) of the proxy-based approaches (f: Köhler; g: Snyder; h: Friedrich) split in time windows with strong or weak divergence of  $\Delta T_g$  and CO<sub>2</sub>. Data are split by the zero line in the standardized ratio  $\Delta T_g / \Delta R_{[CO_2]}$  shown in Figure 3b. White squares are data points which are filtered out in the standardizing of the data, and therefore neither considered in strong or weak divergence part, but which contribute to the fit through all data. In most plots the same  $\Delta R_{[CO_2,LI]}$  from K2015 is plotted, while in (d) CCSM3 is based on  $\Delta R_{[LI]}$  from ICE-5G; in (e) we show  $\Delta R_{[CO_2,LI]}$  as used in CLIMBER and LOVECLIM.

the proxy-based reconstructions than of the model-based simulations. From the three alternative time series, based on different assumptions for the polar amplification factor  $f_{pa}$  in K2015, we use the standard case ( $\Delta T_{g1}$ ), in which  $f_{pa}$  is linearly related to  $\Delta T_{NH}$ . However, our conclusions are not dependent on this choice of  $f_{pa}$  and  $\Delta T_g$  (see the application of the alternative temperature time series in Figure S1 in the supporting information). The fact that three alternative formulations of  $\Delta T_g$  can be connected to the same  $\Delta R_{[LI]}$  shows that there are some degrees of freedom in the connection of both variables.

In K2015 the radiative forcing of  $CO_2$  ( $\Delta R_{[CO_2]} = 5.35 \cdot \ln(CO_2/(278 \text{ ppm})) \text{ W/m}^2$ , Myhre et al. (1998)) and land ice albedo was considered explicitly—leading to  $\Delta R_{[CO_2,LI]}$  and to the state-dependency in  $S_{[CO_2,LI]}$ . It should be noted that when following the revised formulation of Etminan et al. (2016),  $\Delta R_{[CO_2]}$  differs by less than  $0.01 \text{ W/m}^2$  (Köhler, Nehrbass-Ahles, et al., 2017). Furthermore, we assume that radiative forcing is state-independent, which might be a simplification (e.g., Forster et al., 2016). We will analyze similar variables based on alternative  $\Delta T_g$  from proxies (F2016; Snyder, 2016) and simulations (LOVECLIM (F2016), CLIMBER (Ganopolski & Calov, 2011), CCSM3 (He, 2011; Liu et al., 2009)). We will first analyze these different  $\Delta T_g$  in relation to the same  $\Delta R_{[CO_2,LI]}$  as derived in K2015, but for in-depth investigations of simulations we only use the internally applied radiative forcing. The use of these alternative  $\Delta T_g$  for identical  $\Delta R_{[CO_2,LI]}$  has the potential to introduce a bias because temperature and land ice distribution are firmly linked through deconvolution of the LR04 benthic  $\delta^{18}O$  stack. This potential bias is not investigated any further here, although alternative land ice distribution (e.g., ICE-5G of Peltier, 2004) agrees well with our results (K2015). Alternative approaches to estimate  $\Delta R_{[LI]}$  from sea level changes have shortcomings, since they omit the latitudinal effect of land ice distribution on radiative forcing (see K2015 for further details). Chronological misfits between the different records, which might also be introduced in that way, should not be of importance here, as our final interpretations are based on 8-kyr running means. Details of both alternative  $\Delta R_{[LI]}$  estimates and chronological issues have been discussed previously (K2015; Köhler, Stap, et al., 2017). For the CLIMBER simulations additional processes ( $CH_4$ ,  $N_2O$ , vegetation, and aerosols) in the radiative forcing term  $\Delta R_{[X]}$  are also considered.

Time series are standardized before analysis. Due to very high variability in calculated ratios (Figures 3b and 3c, and S1b and S1c) data far away from the mean ( $|\Delta T_g/\Delta R_{[CO_2,LI]}| > 0.25\sigma$ ;  $|\Delta R_{[LI]}/\Delta R_{[CO_2]}| > 1\sigma$ ) are considered as outliers and removed. The chosen cutoff thresholds mainly influence the peak height in the standardized time series, but not the dynamics contained in the time series. Due to the rather linear behavior of the simulations, no outliers in  $\Delta T_g/\Delta R_{[CO_2,LI]}$  have been removed from the LOVECLIM and CLIMBER results. Finally, the outlier-free time series are standardized a second time to enable comparison between the different approaches. This outlier selection during standardization is illustrated for K2015 in Figure S2.

The land ice dynamics simulated in CLIMBER (which are also used in LOVECLIM via off-line coupling) are restricted to Northern Hemisphere ice sheets, Antarctic land ice is kept fixed at present-day configuration, while in K2015 the dynamics of ice sheets and ice shelves in both hemispheres have been investigated. The CCSM3 simulations (He, 2011; Liu et al., 2009) were driven by the ICE-5G land ice distribution, which was compared to de Boer et al. (2014) in K2015. This ICE-5G-based  $\Delta R_{[LI]}$  is also used here when investigating CCSM3 results.

We use the internal fitting routines of the software package GLE, the Graphics Layout Engine (<http://www.gle-graphics.org>) and use  $F$  tests to determine whether a second-order polynomial fits the scattered  $\Delta T_g - \Delta R$  data better than a linear approach (Table S1). For all fits the precondition of meeting the origin is applied (no temperature change for no forcing change), leading to the following two regression equations to be tested: either  $y = b \cdot x$  (linear) or  $y = b \cdot x + c \cdot x^2$  (nonlinear).

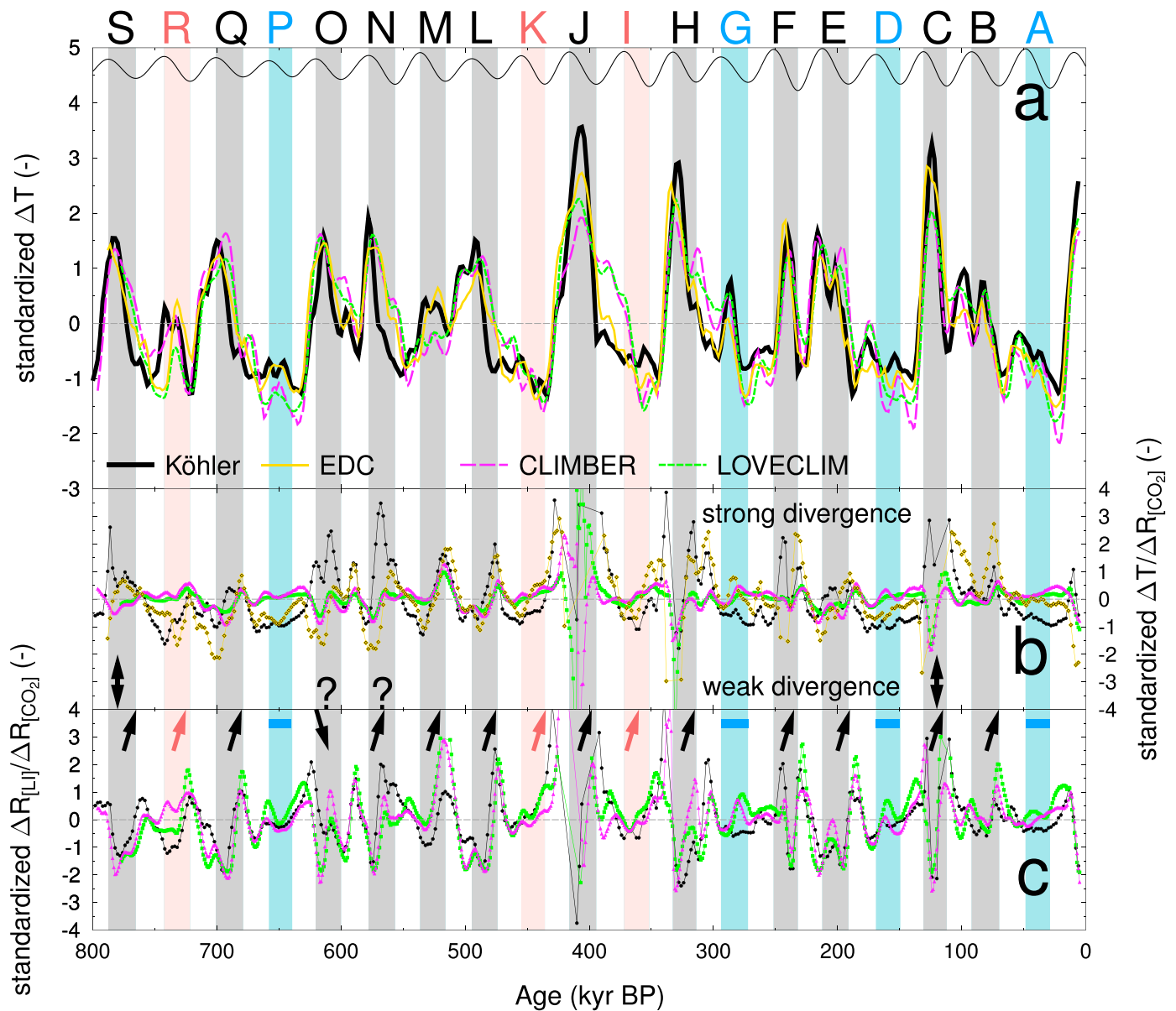
In cases where uncertainties in both  $\Delta T_g$  and  $\Delta R_{[X]}$  are available, more elaborate statistics might be applied (e.g., Monte Carlo approaches have been used in K2015). Uncertainties in  $\Delta T_g$  are only available for K2015 and Snyder. In Figure S3, we show that nonlinear fits are very similar when considering or ignoring uncertainties in these two data sets. We take this as support for the more simplistic approach in our main analysis: all data sets are treated identically and fits are calculated without considering uncertainties in the scattered data.

### 3. Results and Discussions

#### 3.1. Proxy-Based Reconstructions Versus Model-Based Simulations

The main difference between proxy-based reconstructions and model-based simulations to estimate global temperature changes is that the proxy-based reconstructions capture the impacts of all Earth system processes active in the considered time window, while in the model-based approaches only those processes





**Figure 3.** Multimillennial (all data as 8-kyr running mean)  $\Delta T$ - $\text{CO}_2$  divergence and relative contributions of radiative forcing of land ice albedo and  $\text{CO}_2$  for  $\Delta T$  in different setups. (a)  $\Delta T$  (local  $\Delta T$  for EDC and  $\Delta T_g$  elsewhere); (b) the divergence of  $\Delta T$  and  $\text{CO}_2$  described by  $\Delta T/\Delta R_{[\text{CO}_2]}$ ; (c)  $\Delta R_{[\text{LI}]}/\Delta R_{[\text{CO}_2]}$ : Relative land ice (sea level) contribution with respect to  $\text{CO}_2$ . The data sets Köhler and EDC differ only by their  $\Delta T$ . From the model simulations (CLIMBER and LOVECLIM) we analyzed the internally used radiative forcing. All data sets have been standardized and outliers in the ratios have been filtered out. Obliquity (Laskar et al., 2004) is sketched on top of subpanel a (thin black line), with shadings and labels (A–S) indicating times of decreasing obliquity. Color code is given by the details of the Köhler data set: Gray = strong  $\Delta T_g$ - $\text{CO}_2$  divergence including large variations in relative sea level contribution; light red = no or weak  $\Delta T_g$ - $\text{CO}_2$  divergence and large variations in relative sea level contribution; light blue = no or weak  $\Delta T_g$ - $\text{CO}_2$  divergence and stable relative sea level contribution. Vertical two-headed arrows in the  $\Delta T_g$ - $\text{CO}_2$  divergence panel indicate the antiphase dynamics partially seen between Köhler and the CLIMBER/LOVECLIM data sets. Question marks in (b) highlight two phases (MIS 15a and MIS 15e) during which Köhler and EDC largely disagree.

implemented in the model can leave their imprint in the simulation results. Simulated time series of  $\Delta T_g$ , therefore, have to be questioned critically for any serious omissions. In other words, any persisting difference between proxy-based reconstructions and simulated  $\Delta T_g$  might be caused by those processes not included in the models. Alternatively, proxy-based reconstructions might be systematically biased, although this seems unlikely if independent reconstructions come to similar conclusions.

Here we compare results of others to the approach of K2015 (Figure 1a) in order to understand when the proposed state-dependency in  $S_{[\text{CO}_2, \text{LI}]}$  is sustained or when it needs to be rejected. If we replace  $\Delta T_g$  with

an alternative time series (F2016, Snyder, CLIMBER, LOVECLIM, CCSM3, Figures 1b and 1c), we find a similar state-dependency in  $S_{[\text{CO}_2, \text{LI}]}$ —with higher values for warmer conditions—when the applied  $\Delta T_g$  time series is based on proxy-based reconstructions (Figure 2b). This holds for the temperature data set of Snyder, as well as for proxy-based  $\Delta T_g$  derived in F2016 (Figure 2b). The nonlinearity in the  $\Delta T_g$ - $\Delta R_{[\text{CO}_2, \text{LI}]}$  scatter plots is less pronounced in these alternative calculations, when compared to K2015.

If temperature anomalies are taken from CLIMBER simulations, a nonlinear relationship between  $\Delta T_g$  and  $\Delta R_{[\text{CO}_2, \text{LI}]}$  is generated that is inverse to that found by K2015 (Figure 2c), suggesting a smaller paleoclimate sensitivity for warmer climates. Similarly, if we base this analysis on the  $\Delta T_g$  simulated in LOVECLIM, we find an inverse nonlinear relationship—opposite to the proxy-based results (Figure 2c). Since the  $\Delta T_g$ - $\Delta R_{[\text{CO}_2, \text{LI}]}$  relationship of the proxy-based reconstructions of F2016 and the transient LOVECLIM simulations show the opposite slope, it is natural that an averaged  $\Delta T_g$  based on both (as used in F2016) contains a rather linear relationship (Figure 2d). Finally, we analyzed the only available transient GCM-simulation, the Trace21K scenario of the CCSM3 model for the last 21 kyr. Using their  $\Delta T_g$ , we again find the same results as from the EMIC runs (Figure 2d)—a state-dependent paleoclimate sensitivity with steeper slopes in the  $\Delta T_g$ - $\Delta R_{[\text{CO}_2, \text{LI}]}$  data during colder climates, pointing to a higher  $S_{[\text{CO}_2, \text{LI}]}$ , which is inverse to the results from the proxy-based approaches.

If we analyze internally consistent EMIC simulation results using the radiative forcing of  $\text{CO}_2$  and land ice applied in the model runs together with the simulated  $\Delta T_g$  (instead of  $\Delta R_{[\text{CO}_2, \text{LI}]}$  based on K2015), we find a linear relationship between  $\Delta T_g$  and  $\Delta R_{[\text{CO}_2, \text{LI}]}$  for LOVECLIM (Figure 2e). In CLIMBER we find a similar nonlinear relationship between  $\Delta T_g$  and  $\Delta R_{[\text{CO}_2, \text{LI}]}$ —with steeper slope during cold climate—as in the approaches in which the CLIMBER-simulated  $\Delta T_g$  was analyzed together with  $\Delta R_{[\text{CO}_2, \text{LI}]}$  of K2015 (Figure 2e). Further details on the differences in  $\Delta R_{[\text{LI}]}$  for the different approaches can be found in Figure S4.

### 3.2. Obliquity-Driven Changes and the $\Delta T_g$ - $\text{CO}_2$ Relationship

How can we understand this strong state-dependency of  $S$  found in proxy-based approaches and the difference to the model-based approaches? It has recently been deduced, from ice core data covering the last 800 kyr, that the multimillennial trend of atmospheric  $\text{CO}_2$  concentration and Antarctic temperature diverge when obliquity decreases (Hasenclever et al., 2017). One way of perceiving this divergence is that the reduced incoming insolation at high latitudes causes land ice sheet growth and cooling, while there is a coexisting process that keeps  $\text{CO}_2$  at a relatively constant level. Solid Earth modeling experiments have indicated that falling sea level might lead to enhanced magma and  $\text{CO}_2$  production at mid-ocean ridges (e.g., Lund & Asimow, 2011). Hasenclever et al. (2017) suggested that the combination of marine volcanism at mid-ocean ridges and at hot spot island volcanoes might react to decreasing sea level and be a potential cause for this  $\Delta T_g$ - $\text{CO}_2$  divergence. Alternatively, the divergence implies that processes other than  $\text{CO}_2$  radiative forcing or land ice albedo (potentially radiative forcing from non- $\text{CO}_2$  GHGs, or albedo change caused by aerosols, or vegetation) dominate during these phases—leading to a cooling with little reduction in  $\text{CO}_2$ . The evidence so far (e.g., Köhler et al., 2010) does not indicate that the latter was the case, although potential impacts of different forcing efficacy (e.g., Stap et al., 2018; Yoshimori et al., 2011) have so far not been investigated. One study analyzed the contribution of the terrestrial carbon cycle to the divergence of  $\text{CO}_2$  and  $\Delta T_g$  at the end of the present (Holocene) and the previous (Eemian or MIS 5e) interglacial (Brovkin et al., 2016). Processes which seemed to explain the reconstructed divergence in the Holocene failed to explain similar dynamics during MIS 5e, pointing to model deficiencies in the representation of the land carbon cycle, or suggesting that other processes are at work. All modeling results used in here (CLIMBER, LOVECLIM, and CCSM3) were obtained in simulations with prescribed observed  $\text{CO}_2$  concentrations and thus include all effects of the Earth system feedbacks on  $\text{CO}_2$ . However, simulation results do not contain the characteristic long-term  $\Delta T_g$ - $\text{CO}_2$  divergence found in the proxy-based reconstructions (Snyder, F2016), or in the deconvolution of LR04- $\delta^{18}\text{O}$  into land ice dynamics (K2015). This suggests that a relatively low rate of simulated land ice growth and associated cooling during times of decreasing obliquity, and not a feedback on  $\text{CO}_2$ , might be responsible for the difference between model- and proxy-based approaches.

When  $\Delta T_g$  is derived mainly from proxy-based reconstructions (K2015, F2016, and Snyder), our results show a strong  $\Delta T_g$ - $\text{CO}_2$  divergence at times of obliquity decrease. An example of this is the dynamics at the end of the Eemian (see zoom-in in the inset in Figure 1a). For comparison of the different approaches, all time series in the following are analyzed in their standardized versions (Figures 3 and S1). They confirm the earlier finding of a temperature- $\text{CO}_2$  divergence at times of obliquity decrease by Hasenclever et al. (2017), in which not

global temperature change, but Antarctic temperature change derived from the EPICA Dome C (EDC) ice core (Jouzel et al., 2007) has been considered. The temporal evolution of this divergence between  $\Delta T_g$  and  $\text{CO}_2$  can be observed by analyzing the multimillennial dynamics of the ratio  $\Delta T_g / \Delta R_{[\text{CO}_2]}$ , which by coincidence is also defined as  $S_{[\text{CO}_2]}$  (Figure 3b). The interpretation of  $S_{[\text{CO}_2]}$  as a proxy for the multimillennial  $\Delta T_g$ - $\text{CO}_2$ -divergence represents a major improvement in the understanding of  $S_{[\text{CO}_2]}$ , since previously no meaningful patterns have been detected in its temporal variability (PALAEOSSENS-Project Members, 2012). We find that a strong  $\Delta T_g$ - $\text{CO}_2$  divergence exists in 12 out of 19 phases with decreasing obliquity (gray bands in Figure 3) in the data from K2015. Furthermore, the ratio of land ice and  $\text{CO}_2$  radiative forcing ( $\Delta R_{[\text{LI}]} / \Delta R_{[\text{CO}_2]}$ ) underwent large changes during these intervals (Figure 3c), suggesting that land ice (sea level) related changes might indeed be connected to the times of these diverging trends.

The seven phases with decreasing obliquity, but without strong  $\Delta T_g$ - $\text{CO}_2$  divergence in K2015, can furthermore be divided into periods with a stable ratio of  $\Delta R_{[\text{LI}]} / \Delta R_{[\text{CO}_2]}$  (light blue bands marked A, D, G, and P) and those with strong variability in  $\Delta R_{[\text{LI}]} / \Delta R_{[\text{CO}_2]}$  (light red bands I, K, and R). In the former periods (blue colored) the stable ratio of land ice and  $\text{CO}_2$  radiative forcing suggests in-phase variations of both processes, which might indicate that any potential sea level-related  $\text{CO}_2$  outgassing from marine volcanism or other processes could be compensated by the land ice sheet albedo feedback. In the latter periods (red colored) the ratio  $\Delta T_g / \Delta R_{[\text{CO}_2]}$  is always increasing toward the end of the obliquity-half cycle, suggesting that some sea level-related process affecting  $\text{CO}_2$  might have initiated but not yet developed its full potential. This leads, for example, to the unusual strong  $\Delta T_g$ - $\text{CO}_2$  divergence after the end of period K at 436 kyr BP which persisted for almost a complete obliquity cycle around MIS 11. Five of these seven phases with decreasing obliquity but without a strong  $\Delta T_g$ - $\text{CO}_2$  divergence (A, D, I, K, and P, but not G and R) are also characterized by very modest cooling, indicating that the net climate changes during these phases are small when compared to other phases with decreasing obliquity. These phases should, therefore, be interpreted with care since the dominant climate variations occur during other times.

Much smaller variations in the  $\Delta T_g$ - $\text{CO}_2$  divergence are found when analyzing model-based simulations of CLIMBER and LOVECLIM than in K2015 (Figure 3b). Furthermore, the model-based  $\Delta T_g$ - $\text{CO}_2$  divergence observed during times of decreasing obliquity is partially in antiphase to the proxy-based results (phases C and S), suggesting highly synchronous variations in  $\text{CO}_2$  and simulated  $\Delta T_g$  while a strong divergence to  $\text{CO}_2$  persists in the reconstructed  $\Delta T_g$  (Figure 3b). The two lukewarm interglacials MIS 15a, and 15e (phases N, O, 570 and 610 kyr BP, respectively, Past Interglacials Working Group of PAGES, 2016) seem to be special in this respect, since the  $\Delta T_g$ - $\text{CO}_2$  divergence from K2015 is in antiphase to those based on the simulation output and also to that based on EDC  $\Delta T$ . Interestingly, the temperature- $\text{CO}_2$  divergence during the MIS 5/4 transition, around 75 kyr BP (phase B) which motivated the study of Hasenclever et al. (2017), is one of the largest in EDC but rather weak in K2015. Our calculated  $\Delta T_g$ - $\text{CO}_2$  divergence, based on  $\Delta T_g$  of Snyder or F2016, contains qualitatively similar dynamics related to obliquity as that based on EDC  $\Delta T$  or K2015 but differs from the model-based simulations (Figure S1). This qualitative agreement of the divergence in proxy-based  $\Delta T_g$  (K2015, F2016, Snyder, and EDC) provides confidence in the global temperature record obtained in K2015. Furthermore, tests have shown that if new insights into polar amplification (Stap et al., 2018) are used for an improvement of the model setup used in K2015, only small changes in  $\Delta T_g$  are generated, but the general difference to the model-based simulations persists. Based on these findings, the analysis of Hasenclever et al. (2017) needs to be expanded: decreasing obliquity seems to be a necessary but not a sufficient condition for the  $\Delta T_g$ - $\text{CO}_2$  divergence. Another process related to sea level change, or in detail to  $\Delta R_{[\text{LI}]} / \Delta R_{[\text{CO}_2]}$ , needs to be active at the same time to explain the data.

The importance of this  $\Delta T_g$ - $\text{CO}_2$  divergence and its connection to obliquity, for the state-dependency of our paleoclimate sensitivity estimate, becomes apparent when we split the data into times with increasing or decreasing obliquity. In the latter case the nonlinearity (parameter  $c$  in the second-order fit) between  $\Delta T_g$  and  $\Delta R$  is significantly different in the data set of K2015 and Snyder (Figures S5a and S5c), while in the CLIMBER output hardly any difference can be detected (Figure S5b). For F2016 (Figure S5d), which shows a nonlinear relationship when all data are analyzed, the relationship is only linear in both data subsets when differentiated by their phase of obliquity. When data are split based on the ratio  $\Delta T_g / \Delta R_{[\text{CO}_2]}$  in subsets with strong or weak  $\Delta T_g$ - $\text{CO}_2$  divergence, we find an even larger difference in the nonlinearity than when data are split by obliquity in K2015 (Figure 2f), implying a more linear relationship for data with strong  $\Delta T_g$ - $\text{CO}_2$  divergence than for data with decreasing obliquity. When using  $\Delta T_g$  from the proxy-based reconstructions of Snyder and F2016, we find a nonlinear relationship in the  $\Delta T_g$ - $\Delta R_{[\text{CO}_2, \text{LI}]}$  scatter plot during strong  $\Delta T_g$ - $\text{CO}_2$  divergence, while for



times with more synchronous changes in  $\Delta T_g$  and  $\text{CO}_2$  (weak divergence) a linear relationship between  $\Delta T_g$  and  $\Delta R_{[\text{CO}_2, \text{LI}]}$  emerges (Figures 2g and 2h).

### 3.3. Using Paleoclimate Sensitivity to Estimate $\Delta T_{2\times\text{CO}_2}$

The  $\Delta T_g$ - $\text{CO}_2$  divergence appears mainly during, or in connection with, periods of decreasing obliquity related to land ice growth or sea level fall. These times cover  $\sim 50\%$  of past climates. We conclude that for a generic climate system understanding the implementation of the processes responsible for this  $\Delta T_g$ - $\text{CO}_2$  divergence, potentially being the solid Earth-climate feedbacks related to a sea level-induced change in marine volcanism (e.g., Hasenclever et al., 2017; Lund & Asimow, 2011), is essential.

Intervals of strong  $\Delta T_g$ - $\text{CO}_2$  divergence should not be considered for the interpretation of paleodata in the context of future warming, for example, by calculating the paleoclimate sensitivity  $S$ , because in the future we expect sea level to rise. Otherwise the climate system response of a glaciation is erroneously implicated with anthropogenic warming. Here one might rely only on the subset of  $\Delta T_g$ - $\Delta R$  data that coincide with times of weak (or no)  $\Delta T_g$ - $\text{CO}_2$  divergence. For K2015, this restriction would lead to a different quantification of paleoclimate sensitivity following the framework of Köhler, Stap, et al. (2017; Figure 2f). In detail,  $S_{[\text{CO}_2, \text{LI}]}$  can be derived from the fit to the scattered  $\Delta T_g$ - $\Delta R_{[\text{CO}_2, \text{LI}]}$  data after  $S_{[\text{CO}_2, \text{LI}]} = b + c \cdot \Delta R_{[\text{CO}_2, \text{LI}]}$ . The paleodata of the last 800 kyr cover mainly intervals with  $\Delta R_{[\text{CO}_2, \text{LI}]} \leq 0 \text{ W/m}^2$ , and due to the state-dependent character of  $S_{[\text{CO}_2, \text{LI}]}$  we refrain from an extrapolation of our derived fitting function to a range not covered by the data, for example, to  $\Delta R_{[\text{CO}_2, \text{LI}]} > 0 \text{ W/m}^2$ . Nevertheless, climates comparable to late Pleistocene interglacials can be approximated by  $\Delta R_{[\text{CO}_2, \text{LI}]} \approx 0 \text{ W/m}^2$ .  $S_{[\text{CO}_2, \text{LI}]}$  for those interglacials would be  $\sim 20\%$  smaller when excluding intervals of  $\Delta T_g$ - $\text{CO}_2$  divergence in comparison to calculations based on all available data,  $S_{[\text{CO}_2, \text{LI}]} = 1.6 \text{ K/(W/m}^2\text{)}$  instead of  $2.0 \text{ K/(W/m}^2\text{)}$ . If based on  $\Delta T_g$  of Snyder (Figure 2g) or F2016 (Figure 2h) these subsets of data with weak (or no)  $\Delta T_g$ - $\text{CO}_2$  divergence are defined by a linear relationship between  $\Delta T_g$  and  $\Delta R_{[\text{CO}_2, \text{LI}]}$  and a constant  $S_{[\text{CO}_2, \text{LI}]}$  of  $0.82$  and  $0.88 \text{ K/(W/m}^2\text{)}$ , respectively. To estimate equilibrium warming caused by  $2\times\text{CO}_2$  ( $\Delta T_{2\times\text{CO}_2}$ , the classical Charney ECS (Charney et al., 1979; Knutti et al., 2017) from our  $S_{[\text{CO}_2, \text{LI}]}$ , we need to correct for missing slow processes (radiative forcing of  $\text{CH}_4$  and  $\text{N}_2\text{O}$ ; albedo changes caused by vegetation and aerosols). In a previous study (PALAEOSSENS-Project Members, 2012) the ratio between  $S_{[\text{GHG, LI, VG, AE}]} / S_{[\text{CO}_2, \text{LI}]}$  for the last 800 kyr has been determined as  $0.64 \pm 0.07 (1\sigma)$ . Note that this correction for the slow processes ignores any state-dependency that might be associated with them. Together with the average radiative forcing for a doubling of  $\text{CO}_2$  of  $3.71 \text{ W/m}^2 (\pm 10\% (1\sigma))$  (Myhre et al., 1998) our  $S_{[\text{CO}_2, \text{LI}]}$  for late Pleistocene interglacials translates into a  $\Delta T_{2\times\text{CO}_2}$  or ECS of  $1.9 \pm 0.3 \text{ K}$  (Snyder),  $2.1 \pm 0.3 \text{ K}$  (F2016), and  $3.8 \pm 0.6 \text{ K}$  (K2015). Alternative calculations, based on the data split by obliquity (Figure S5), would lead to slightly larger numbers of ECS ( $2.3 \pm 0.3 \text{ K}$  (Snyder),  $2.3 \pm 0.3 \text{ K}$  (F2016), and  $4.4 \pm 0.7 \text{ K}$  (K2015)); however, we consider these to be less reliable following our analysis in the previous subsection. This compares well with other approaches (Knutti et al., 2017), including the narrow *likely* (66% confidence interval) range of  $2.2$ – $3.4 \text{ K}$  recently obtained from an emerging constraint from global temperature variability and CMIP5 (Cox et al., 2018), and the 95% confidence range of  $2.0$ – $4.3 \text{ K}$  from a large model ensemble, which has been constrained by observational and geological evidences (Goodwin et al., 2018).

## 4. Conclusions

In conclusion, we find an inconsistency in the state-dependency of paleoclimate sensitivity calculated from model simulations and proxy-reconstructions, when explicitly considering radiative forcing of  $\text{CO}_2$  change and land ice albedo change, or  $S_{[\text{CO}_2, \text{LI}]}$ . This may be related to the fact that fast climate feedbacks in EMICs are too linear. Furthermore, EMICs may underestimate the strength of some slow climate feedbacks. As it has been shown that solid Earth-climate feedbacks can play an important role for  $\text{CO}_2$  dynamics during glacial cycles (e.g., Hasenclever et al., 2017; Huybers & Langmuir, 2009; Lund & Asimow, 2011), these feedbacks should be incorporated in models used to simulate  $\text{CO}_2$  concentration (e.g., Ganopolski & Brovkin, 2017). Furthermore, one also needs to fully understand why current model simulations contain none of the temperature- $\text{CO}_2$  divergence observed during intervals of decreasing obliquity within proxy-based reconstructions. Our study suggests that one possible reason for this discrepancy is that the CLIMBER model underestimates the rate of land ice growth during periods of decreasing obliquity and consequently simulates less cooling induced by land ice. It should be emphasized that the magnitude of the expected  $\text{CO}_2$  changes connected with these solid Earth feedbacks are small when compared with anthropogenic  $\text{CO}_2$  changes. Therefore, these missing model feedbacks in CLIMBER do not affect its ability to simulate future temperature increase caused by a rise

in CO<sub>2</sub>. Our results have important consequences for future efforts to quantify paleoclimate sensitivity from proxy-based analyses. We suggest that studies should focus on intervals without decreasing obliquity or sea level, since the detected divergence of global temperature and CO<sub>2</sub> during these intervals could otherwise overprint the system response.

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